Seismological Institute
University
Uppsala, Sweden

# SEISMIC INVESTIGATION OF CRUSTAL STRUCTURE

Contractor: Seismological Institute, University, Uppsala, Sweden

Contractor's Annual Technical Report Number 1

Covering the period November 1, 1960 - October 31, 1961

| | |
|---|---|
| Contract Number | DA-91-591-EUC-1637 |
| Date of Report | November 3, 1961 |

Prepared for

EUROPEAN RESEARCH OFFICE
U.S. DEPARTMENT OF THE ARMY
Frankfurt/Main, Germany
APO 757, U.S. Forces

Work sponsored by Advanced Research Projects Agency
Project VELA UNIFORM
ARPA Order No. 165-61



Markus Båth
Markus Båth
Project Scientist

# SEISMIC INVESTIGATION OF CRUSTAL STRUCTURE

## Abstract


The research has closely followed the lines which were drawn up in the Research Proposal, dated June 28, 1960. Six papers have been prepared for publication, comprising three main items:

1) Seismic investigation of crustal structure in Iceland and the North Atlantic region.
2) Seismic investigation of crustal structure in Fennoscandia.
3) Studies of amplitudes of explosion-generated seismic waves.

1) The records obtained by refraction technique during a Swedish-Icelandic seismic expedition to Iceland in 1960 have been analysed. The detailed structure of the lava layer down to about 5 km depth has been determined along a profile across Iceland from southwest to northeast. The structure was found to be more complex than expected from earlier results. This determination is the first more detailed investigation of lava structure, at least on Iceland. This research constitutes our Scientific Report No. 1 and has been published in the Journ. Geophys. Res., Vol. 66, No. 6, 1961, pp. 1913-1925.

The structural relation between Iceland and the surrounding ocean, especially the Mid-Atlantic Ridge, has been investigated by means of group-velocity dispersion curves for Love and Rayleigh waves from earthquakes in the region. This research constitutes our Scientific Report No. 3 and has been submitted for publication in the Bull. Seism. Soc. Amer.

Investigations of travel times of P waves from earthquakes in the Arctic-Atlantic Ocean have given results with regard to the deeper structure in the same region. Thus, it has been demonstrated that the P wave velocity in the upper mantle below the Arctic-Atlantic Ocean is around 7.4 km/sec,

while below northwest Europe it is around 8.36 km/sec. In the upper mantle below the ocean area mentioned, it appears to be a discontinuity at about 140 km depth, where the P wave velocity rapidly increases to about 8.2 km/sec. There are indications of low-velocity layers in the upper mantle both below the ocean and below the continent. This research has appeared as our Scientific Report No. 5 and has been submitted for publication in Annali di Geofisica.

2) The thickness of the crust in Fennoscandia has been determined by the phase velocity method, applied to Rayleigh waves from two earthquakes (Mexico and Kurile Islands) recorded by Swedish, Danish and Finnish stations. Large distances between stations and inhomogeneous structure in the neighbouring areas made a straight-forward application of the method very difficult. Therefore, improved methods were developed to take account of such difficulties. The crustal thickness was close to 35 km for the whole area investigated, with indication of slightly thicker crust in northern Sweden (35-38 km) as compared to Finland and southern Sweden (33-35 km). This research has appeared in our Scientific Report No. 2 and will be published in Annali di Geofisica.

The records obtained from a seismic reflection experiment at Kiruna in 1955 have been analysed. This was the first seismic investigation of the deeper structure undertaken in Fennoscandia. Direct P and S waves as well as atmospheric sound waves were recorded. The reflected signals were generally weak and of erratic occurrence. Approx. depths to the Conrad and the Mohorovičić discontinuities were calculated as 19 km and 33-34 km resp. This research has been presented in our Scientific Report No. 4 and will be published in Geofisica pura e applicata.

3) Although not explicitly contained in the contract, we have also studied the amplitudes obtained in the seismic refraction records in Iceland

in 1959 and 1960. These have permitted to establish relations between amplitudes and charge weights, furthermore to study the effect of water cover for underwater explosions, and finally to determine the attenuation of P waves in Iceland. This research constitutes the content of our Scientific Report No. 6 and has been submitted for publication in Geofisica pura e applicata.

## 1. Definitions

The word crust has a meaning of a hard surface layer, covering softer interior. The earth's crust originally indicated the solid surface layers, covering the interior of the earth, which was believed to consist of fluid or viscous material. Later, different geophysical measurements showed that the earth behaved as solid material in relation to seismic waves to a depth of 2900 km below the surface. However, the word crust has not been applied to this 2900 km thick shell, but to represent only a relatively thin superficial layer. The lower boundary of the crust has been subject to several definitions, as the boundary between crystalline and vitreous state of the material, boundary between the lithosphere (outer shell having a yield strength of the order of $10^9$ dynes/$cm^2$) and the asthenosphere (with much smaller yield strength) or even the boundary between the roughly 700 km thick shell in which earthquakes occur and the deeper parts of the earth, where earthquakes are not known to occur (Gutenberg 1959). However, most geophysicists adopt the definition that the Mohorovičić discontinuity, hereafter called the Moho, separates the crust from the mantle. This definition will be followed here.

Apparently, there exists no exact and generally accepted definition of the Moho. In most cases it is sufficient to define this as a velocity discontinuity near the surface of the earth, where longitudinal wave velocity jumps to about 8 km/sec.

As the crust covers the whole earth, and is separated from the mantle by the Moho, according to the adopted definition, the definition of Moho must be such, that it exists below any point on the earth's surface. This is fulfilled by the definition, that the Moho is taken at the depth, where the compressional wave velocity first exceeds 7.8 km/sec (*Steinhart* and *Meyer* 1961). This definition does not specify any discontinuity in velocities, and has therefore a physical meaning quite different from that originally associated to the Mohorovičić discontinuity.

We prefer such a definition that Moho always represents a velocity discontinuity, if this exists, or else a level where the velocity increase with depth (velocity gradient) reaches a maximum inside a specified velocity and depth interval. Such a definition may be expressed as follows. *The Moho is the lowest velocity discontinuity in the uppermost 100 km of the earth, where the compressional wave velocity increases to a value between 7.0 and 8.5 km/sec. If no such discontinuity exists, the Moho is at that depth, where the velocity gradient is steepest inside the depth and velocity limits adopted.*

The continental crust is frequently divided into three sections, the *sedimentary* layers, the *granitic* layer and the *basaltic* or *gabbro* layer. The *sedimentary* layers cover practically the whole earth, both continents and ocean bottoms. Their thickness and physical properties vary very much from one place to another, so any definition is difficult. The *granitic* layer is generally reported by seismologists to have a longitudinal wave velocity somewhere between 5.5 and 6.5 km/sec (*Gutenberg* 1959, *Steinhart* and Meyer 1961, Båth 1961 a). Sawarenski and Kirnos (1960, p. 227) define the granitic layer as crustal layers, having longitudinal wave velocity between 5.2 and 6.0 km/sec. This layer is generally not observed below deep oceans. The basaltic or gabbro layer, separated from the granitic layer by the Conrad discontinuity, is frequently reported by seismologists, but its wave velocity seems to vary very much. This is partly due to different definitions adopted, but also due to great variation inside this layer or layers. This layer is found in both continental and oceanic regions. We suggest the following definitions for use in seismological works:

Sedimentary layer: Longitudinal wave velocity lower than 5.2 km/sec.

Granitic layer: Longitudinal wave velocity higher than 5.2 km/sec but lower than 6.2 km/sec.

Basaltic layer: Longitudinal wave velocity higher than 6.2 km/sec but lower than in the upper mantle, usually lower than 7.5 km/sec.

These definitions are arbitrary, but in close agreement with reported velocity of crustal P-waves. The names of the layers do not necessarily indicate the geological composition of the layers, as the definitions are based entirely on wave velocities.

On the Mid-Atlantic Ridge and in Iceland (and also in several other regions) the upper crustal layers consist of volcanic material, erupted during the Tertiary and Quaternary epochs. These layers are here named the lava layer, defined as a surface or near-surface layer in Iceland and on the Mid-Atlantic Ridge, having longitudinal wave velocity higher than 3.5 km/sec but lower than 6 km/sec.

## 2. Seismic Methods for Determining the Structure of the Crust

The propagation of elastic waves through the earth has contributed to our knowledge of the crustal structure and the structure of the earth's interior, more than all other research methods together. The elastic waves are produced either by natural earthquakes or by artificial explosions. The earthquake waves have their advantage in the large energy and amplitudes, but their disadvantage lies in the unknown quantities, the origin time, epicenter and focal depth. These quantities are usually known in artificial explosions.

If natural earthquakes are used for determination of crustal structure, the location of focus and origin time must be determined with the highest possible accuracy. The methods of epicenter location are described in seismological textbooks (e.g. Richter 1958). For well recorded earthquakes an accuracy in epicenter of about 10 km is possible if all circumstances are favourable, by using only stations at more than 2000 km distance. If nearer stations are used, a knowledge of local anomalies in crustal and upper mantle wave velocities is needed, especially if the epicenter is located near a boundary between different structures. The possible accuracy in determined hypocentral depth is difficult to state. In large earthquakes, where PKP is clearly recorded at a number of stations, the hypocentral depth can be determined with great accuracy (possibly 10 km), but in smaller earthquakes, originating at great distance from a dense net of seismograph stations, the accuracy in depth determination is poor (some 50 km).

The seismic methods for investigation of crustal structure can be classified as follows (Båth, 1958):

A: Propagation of body waves.

$A_1$ Refraction using natural earthquakes

$A_2$ Refraction using artificial explosions

$A_3$ Reflection using natural earthquakes

$A_4$ Reflection using artificial explosions

B: Propagation of surface waves.

$B_1$ Phase propagation of Rayleigh waves

$B_2$ Group propagation of Rayleigh waves

$B_3$ Phase propagation of Love waves

$B_4$ Group propagation of Love waves

C: Other seismic methods.

Each of these methods will be described briefly. For mathematical treatment of the methods the following symbols are used:

| | |
|---|---|
| $c$ | Phase velocity |
| $U$ | Group velocity |
| $\alpha$ | Longitudinal wave velocity |
| $\beta$ | Shear wave velocity |
| $L$ | Wave length |
| $T$ | Wave period |
| $t$ | Time, travel time |
| $H$ | Thickness of layer |
| $h$ | Hypocentral depth |
| $\Delta$ | Epicentral distance |
| $\omega$ | Inclination of interface |
| $\mu$ | Rigidity |
| $k$ | $2\pi/cT$ |
| $C$ | Constant |

Indices 1,2,3.....i.....N refer to crustal layers, counted from the earth's surface.

<u>A 1. Refraction using natural earthquakes</u>. The precision of the result depends on the accuracy in epicenter and focal depth determination, together with the number and distribution of near seismograph stations and the accuracy in time keeping at these stations. The observed travel times of P-waves are plotted against epicentral distance. For study of the surface layers we need dense observations up to an epicentral distance some ten times greater than the thickness of the layers to be studied. The observed travel time curve can usually be divided into two or more straight sections, each representing an individual layer, assumed to be homogeneous. For crustal layers, the earth's curvature may be ignored. In a horizontally layered crust, where the focus of the earthquake lies in the surface layer ($h < H_1$), the propagation, along the surface, of waves travelling in this layer is given by the equation

$$t_1 = \frac{\Delta}{\alpha_1} \left[ 1 + \frac{h^2}{\Delta^2} \right]^{\frac{1}{2}} \qquad (1)$$

For waves refracted in the second layer, the travel time equation is

$$t_2 = \frac{\Delta}{\alpha_2} + \frac{2H_1 - h}{\alpha_1} \left[ 1 - \frac{\alpha_1^2}{\alpha_2^2} \right]^{\frac{1}{2}} = \frac{\Delta}{\alpha_2} + C_2 \qquad (2)$$

and for waves refracted in the third layer we get:

$$t_3 = \frac{\Delta}{\alpha_3} + \frac{2H_1 - h}{\alpha_1} \left[ 1 + \frac{\alpha_1^2}{\alpha_3^2} \right]^{\frac{1}{2}} + \frac{2H_2}{\alpha_2} \left[ 1 - \frac{\alpha_2^2}{\alpha_3^2} \right]^{\frac{1}{2}} = \frac{\Delta}{\alpha_3} + C_3 \qquad (3)$$

Generalized to N layers the expression of the travel time becomes

$$t_N = \frac{\Delta}{\alpha_N} + \frac{2H_1 - h}{\alpha_1}\left[1 + \frac{\alpha_1^2}{\alpha_N^2}\right]^{\frac{1}{2}} + 2\sum_{i=2}^{i=N-1} \frac{H_i}{\alpha_i}\left[1 - \frac{\alpha_i^2}{\alpha_N^2}\right]^{\frac{1}{2}} = \frac{\Delta}{\alpha_N} + C_N \qquad (4)$$

If the hypocentral depth is larger than $H_1$, the equations have to be modified (Båth 1960).

The velocities $\alpha_1, \alpha_2, \ldots \alpha_N$ and the time intercepts $C_2, C_3, \ldots C_N$ can be read directly from the travel time curve, and $H_1, H_2, \ldots H_N$ are then easily computed for horizontal layering.

This method has been applied in the investigation reported in Chapter 3.3 below.

A 2. Refraction using artificial explosions. The location of the explosion point and the travel time are usually exactly known. The time keeping is usually much more exact than in case of natural earthquakes, as the recording device is run at much higher speed. Otherwise the treatment and computations are the same as described in A 1. Usually, the refraction method is applied on P-waves only, but in some cases transversal waves (S - waves) may be used to get additional information. If the discontinuity between the first and second layer is dipping and forms the angle $\omega$ with the horizontal plane along the line of measurements, the travel time of the refracted wave is given by the equation

$$t_2 = \frac{1}{\alpha_1}\left[\Delta \sin (i_1 \pm \omega) + 2H_1 \cos i_1 \cos \omega\right] \qquad (5)$$

where

$$\sin i_1 = \alpha_1 / \alpha_2$$

and $H_1$ is the layer thickness under the source.

The minus sign is to be used in direction of decreasing thickness of the surface layer (Gutenberg 1959). If the discontinuity between the lower layers is also dipping, the expressions for the travel times of waves refracted in the deeper layers become more complicated, than in (2), (3) and (4).

This method has been applied in the investigation reported in Chapter 3.1.

A 3. Reflection using natural earthquakes. Reflections from the bottom of the crust or from discontinuities inside the crust are difficult to identify in seismograms of natural earthquakes. This is mainly due to the relatively large distances between the seismograph stations, compared with the layer thicknesses. However, several authors have identified such reflections (e.g. Gutenberg 1944). Because of the difficulty in identification, reflections using natural earthquakes are of limited importance in crustal structure research.

A 4. Reflection using artificial explosions. This method is very important in geophysical prospecting, but there are grave difficulties in using it to investigate the deeper crustal structures. In geophysical prospecting, the geophones are usually placed in a straight line, and an explosive charge is detonated in the midpoint of this line. The impulses from the geophones are fed through amplifiers of special design, including band-pass filters and automatic gain control. By this means, impulses of very small energy or small ground amplitudes may be detected on the seismograms. The method is described in textbooks in geophysical prospecting (e.g. Ewing and Press 1956). For deep reflections, the distance between shot point and geophones may be chosen so that total reflection occurs. The mathematical treatment of such cases is described by Båth (1960).

This method has been applied in the investigation reported in Chapter 3.5. below.

B 1. Phase propagation of Rayleigh waves. The phase velocity of Rayleigh waves in layered media depends on the wave length and the physical constants of the material. The period equation is rather complicated and is given in textbooks on theoretical seismology (e.g. Ewing, Jardetzky and Press 1957). By means of high-speed electronic computers, the period equation can be solved for multilayered structures. A great number of such computations have been made for structures analogous to that of the earth's crust. Observed phase velocities are compared with theoretical ones, and if the velocities are equal for a number of wave frequencies, the crustal structure along the path observed is probably very similar to that used for computation of the theoretical velocities. If only a small part of the wave spectrum is observed, there is a possibility of considerable variation in crustal structure, giving nearly the same velocities in the observed part of the wave spectrum. Furthermore, structural variation along the path may produce the same effect on the observed phase velocity as a crustal structure deviating considerably from that existing.

The actual phase velocity of Rayleigh waves may be obtained by noting the arrival times of wave crests as measured on a homogenoous set of vertical instruments in a dense net of stations. By this procedure, both direction and velocity of each wave crest is obtained, but irregularities in the wave form may cause serious errors, if few stations are used.

This method is applied in the investigation reported in Chapter 3.4.

B 2. Group propagation of Rayleigh waves. If the arrival times of a wave in a dispersed wave train are compared with the origin time of the disturbance and the length of the path traversed, we obtain the group velocity U, which is correlated to the phase velocity c according to the equation

$$U = c + k\frac{dc}{dk} = c - L\frac{dc}{dL} \qquad (7)$$

The group velocity method needs only one station, but the path investigated must be long compared with the wave length. This involves, that the method cannot give information about local variations in the crustal structure, but only the general structure along the whole path. In the application to Rayleigh waves, any possibility of misinterpretation of the wave type must be eliminated. This is most easily done by using only records of vertical seismometers. As there exist several modes of Rayleigh waves (Rayleigh mode, also named first mode, first shear mode or second mode etc) every possibility of taking first shear mode as Rayleigh mode etc must be excluded. Such mistakes are usually easily excluded by comparing the observed group velocities with theoretical ones along structures similar to that expected.

This method is applied in the investigation reported in Chapter 3.2.

B 3. Phase propagation of Love waves. In a layered solid, surface waves with transverse horizontal particle motion exist, called Love waves. Their velocity depends on the wave length and physical constants of the material. For one layer of thickness H over a homogeneous solid half space, the period equation may be written:

$$\tan\left[kH\left(\frac{c^2}{\beta_1^2}-1\right)^{\frac{1}{2}}\right]=\frac{\mu_2}{\mu_1}\,\frac{\left(1-\frac{c^2}{\beta_2^2}\right)^{\frac{1}{2}}}{\left(\frac{c^2}{\beta_1^2}-1\right)^{\frac{1}{2}}} \tag{8}$$

For two solid layers over a homogeneous solid half space, the period equation is somewhat more complicated, but it can easily be solved by using simple hand computers (Ewing, Jardetzky and Press 1957). For more than two surface layers, the period equation is too complicated to be easily solved by using simple computers.

The phase velocity of Love waves can be used for crustal structure research in similar way as phase velocity of Rayleigh waves, but there are some difficulties in selecting pure Love waves on the seismograms. According to our knowledge, this method has never been applied.

B 4. Group propagation of Love waves. The theoretical group velocity of Love waves can be deduced from (7) and (8), or other period equations for Love waves. As Love waves usually travel faster than Rayleigh waves, the beginning of the surface wave train frequently consists of pure Love waves. The group velocity and period of these waves can easily be determined, and used for crustal structure determination in the same way as group velocity of Rayleigh waves. If group velocity of both Rayleigh and Love waves along the same path are used for crustal structure determination, the result will be more reliable, than using only one type of wave.

This method has been applied in the investigation reported in Chapter 3.2.

C. Other application of seismic waves. The amplitudes of body waves reflected at the earth's surface (PP, SS, pP etc) depend somewhat on the crustal structure, especially the layering of the crust in the point of reflection. Some authors have reported double reflections in such cases, one at the Moho, and another at the earth's surface. Time difference between these gives the thickness of the crust. Reflections downwards from the Moho are generally faint and difficult to identify.

Channel waves ($L_g$, $R_g$ etc) are frequently found on seismograms, where the whole path is over continents. Their existence is proof for continental crustal structure along the whole wave path, while their absence does not necessarily indicate oceanic crustal structure (Båth 1961 b).

The attenuation of short period surface waves is found to be much greater in the border zone between continental and oceanic crustal structure, than inside each structure. Furthermore, surface waves of period shorter than 12 sec are usually not found on seismograms of distant earthquakes, where the wave path lies wholly or to a large extent across oceanic crustal structures.

## 3. Results of the Present Research Work

3.1. Upper Crustal Structure of Iceland by Eysteinn Tryggvason and Markus Båth.

In this work the refraction method, using waves from artificial explosions, was applied. Reflections were also used, but these were frequently very vague. The observational data consisted of seismograms obtained in Iceland during the months of August and September 1960, along eight refraction profiles from 20 to 41 km long. This material was used to determine the thickness and wave velocities in the "lava" layer, about 5 km thick. The measuring profiles were laid near a line crossing Iceland from southwest to northeast. The result of this study may be summarized as follows:

1) Surface formations, where P-wave velocity was lower than 3.0 km/sec. This layer consists of irregular surface formations, as sediments of different age, recent lavas and possibly some old lavas. Its thickness is less than 0.2 km in six profiles, but in two profiles in north Iceland, the thickness was found to be 0.35 resp. 0.39 km.

2) Layer with P-wave velocity 3.7 ± 0.3 km/sec. This layer is most pronounced in southwest Iceland, where its thickness is observed to be 1.5 to 2.4 km. In north Iceland the layer is missing. This layer consists probably of irregular lava formations, mainly of Quaternary age. In south Iceland no other lava layer was found.

3) Layer with P-wave velocity 4.95 ± 0.2 km/sec.

This layer lies below layer 2 in central Iceland but below thin sedimentary layer or at the surface in north Iceland. It consists of lava of Tertiary age, where it can be observed at the surface. Its thickness was determined to lie between 1.0 and 2.7 km.

4) Layer with P-wave velocity 5.55 ± 0.05 km/sec.

This layer is found below layer 3 in three profiles in central and northern Iceland. Its thickness is observed to lie between 1.2 and 3.1 km. If this layer is thinner than about 1.0 km, below a layer 3.1 to 2 km thick, it cannot be detected by the refraction method, as no first arrival on the seismograms would correspond to this. The constitution of this layer is unknown, but it is assumed to consist of volcanic material (lava).

5) Layer with P-wave velocity about 6.2 km/sec.

This layer was found below the lava layers at a depth of 1.7 to 3.7 km. At one profile in north Iceland this layer was not found. Its wave velocity is intermediate between the granitic and the gabbro layer, according to the adopted definition, but its composition is unknown.

6) Layer with P-wave velocity about 6.7 km/sec.

This layer is found below the lava layer at 4.8 km depth in one profile in north Iceland. Other observations give its thickness about 15 km (Tryggvason 1959, Båth 1960). It belongs to the gabbro layer according to the definitions adopted.

The result of this research has appeared in our Scientific Report No. 1 and has been published in Journal of Geophysical Research, Volume 66, No. 6, June 1961, pp. 1913-1925.

## 3.2. Crustal Structure of the Iceland Region from Dispersion of Surface Waves

by Eysteinn Tryggvason.

Group velocities of both Love and Rayleigh waves, as determined from records at Icelandic seismograph stations together with one earthquake record from Scoresbysund, are used in this study. The earthquakes studied originated in the Mid-Atlantic Seismic Belt, and the wave paths, 105-1640 km in length, all lie near this seismic belt, and inside the Mid-Atlantic Ridge, if this ridge is defined to include Iceland.

The observed surface wave dispersion was compared with theoretically computed dispersion curves of both Love and Rayleigh waves. The theoretical Love-wave dispersion curves were computed by the writer, while for Rayleigh waves, theoretical curves of Haskell (1953) and Kanai (1951) were used. This comparison showed, that the waves studied were first-mode Love and Rayleigh waves, as otherwise we had to account for a crustal structure, deviating greatly from the previously obtained results from seismic refraction measurements (Ewing and Ewing 1959, Båth 1960). This study shows, that a thin layer of low wave velocity covers the region studied, and controls the surface wave propagation to great extent. The thickness of this surface layer is mainly about 4 km, and therefore the surface waves, which mainly propagate in this layer, are of short period, some 3-6 sec. Waves of longer period are also recorded, especially of earthquakes at considerable distance, but with relatively small amplitudes. For wave paths across Iceland, the Rayleigh-wave train can sometimes be divided into two sections. First arrives a train of small amplitude waves of large period, to be followed by another train of waves of much larger amplitude but shorter period. The dispersion of the first waves is supposed to be caused mainly by surface layers, of total thickness about 10 km or somewhat less. The later, large amplitude waves are also dispersed, but the dispersion is apparently controlled by one surface layer of some 4 km thickness overlying an infinite half-space of homogeneous material. These results lead to the conclusion, that the earth's crust in Iceland can be divided into two sections of similar thickness. Refraction measurements (Tryggvason 1959, Båth 1960) show somewhat different and surely more reliable results about the lower or second crustal layer.

The result of this study is summarized as follows:

The Mid-Atlantic Ridge southwest of Iceland, but north of 52°N. A surface layer about 4 km thick, with S-wave velocity about 2.7 km/sec covers a substratum of unknown thickness, where the S-wave velocity is between 4.0 and 4.5 km/sec, probably about 4.3 km/sec. Nearest Iceland there is an intermediate layer some 5 km thick with S-wave velocity about 3.6 km/sec.

Central Iceland. A surface layer with an S-wave velocity about 2.7 km/sec and some 3 km thick overlies a layer of unknown thickness with S-wave velocity considerably lower than 4.5 km/sec, possibly about 4.0 km/sec. Another possible solution is a surface layer less than 3 km thick with S-wave velocity about 2.5 km/sec overlying a relatively thin ( $<4$ km) intermediate layer of S-wave velocity between 3 and 4 km/sec, which overlies a substratum of higher wave velocity.

Western Iceland. A surface layer 4-5 km thick with S-wave velocity some 2.7 km/sec overlies a layer of similar thickness with S-wave velocity about 3.6 km/sec. The substratum has an S-wave velocity of 4.1 - 4.5 km/sec.

The path SW-Iceland to Scoresbysund. A surface layer of S-wave velocity about 2.7 km/sec and some 7 km thick overlies a substratum of S-wave velocity about 4.3 km/sec.

Together with 3.1 and earlier investigations (Ewing and Ewing 1959, Båth 1960) this study indicates the structural variations from the Mid-Atlantic Ridge to Iceland.

A paper on this study has appeared as our Scientific Report No. 3 and has been submitted for publication in Bulletin of the Seismological Society of America.

3.3. Wave Velocity in the Upper Mantle below the Arctic-Atlantic Ocean and Northwest Europe by Eystoinn Tryggvason.

The refraction method, using waves from natural earthquakes, is applied on data of four earthquakes in the Arctic-Atlantic Ocean. The origin times and epicenters of the earthquakes are as follows:

| | | | | |
|---|---|---|---|---|
| 1951, June 6, | 16 10 47, | 71°.4 N | 10°.1 W |
| 1952, Dec 10, | 05 58 05, | 71°.0 N | 6°.8 W |
| 1958, Jan 23, | 13 35 03, | 65°.2 N | 6°.8 E |
| 1959, Jan 29, | 23 24 29, | 71°.0 N | 7°.2 E |

The epicenters and origin times were determined from arrival times of P-waves at stations at more than 20° distance from the epicenters. No indication was found of abnormal focal depth.

The upper mantle velocity in Fennoscandia was found to be 8.36 km/sec by assuming horizontal layering. This velocity was well defined in the earthquakes of 1958, Jan 23, and 1959, Jan 29. The time intercept for the latter was 11 sec against 7 sec for that of 1958, Jan 23. This difference could possibly be caused by different focal depths, and thus incorrect origin times, but we need a depth difference of some 50 km or more to explain this difference in time intercept. Another explanation, assumed by the writer, is that the difference in time intercept is partly or wholly produced by a low-velocity region in the vicinity of the epicenter of Jan 29, 1959. We have seen (3.2.) that the upper mantle (or lower crustal layers) have a P-wave velocity about 7.4 km/sec in the Iceland region. By assuming that the low-velocity upper mantle forms a broad belt (say 500 km) on both sides of the Atlantic Seismic Belt, we should have the P waves from earthquakes in the Atlantic Seismic Belt somewhat delayed at short distances, while at larger distances the delay is smaller, and depends on the thickness of the mantle layer with abnormally low velocity. If we assume, that the paths from the earthquakes of 1951, 1952, and 1959 to stations in Iceland and to Scoresbysund and Nord in Greenland, lie wholly inside the mantle low-velocity region, we obtain an apparent velocity near 7.5 km/sec up to a distance of about $10^{\circ}$, whereafter the apparent velocity is somewhat in excess of 8 km/sec until the twenty-degree discontinuity is reached, probably at $15^{\circ} - 16^{\circ}$ distance. There is a possibility of a shadow zone at distances between $8^{\circ}$ and $16^{\circ}$, although the earthquake of 1959, Jan 29, was clearly recorded at these distances.

As a preliminary result, the 7.4-layer extends to 100-150 km depth, and thus must belong to the mantle according to the adopted definitions. This may possibly be expressed by saying that the low-velocity asthenosphere layer reaches the top of the mantle in the vicinity of the Mid-Atlantic Seismic Belt.

Below the continent in northwest Europe the P-wave velocity below Moho is found to be unusually high, 8.36 km/sec. A shadow zone is not pronounced as reported from some other regions (Gutenberg 1959). We have clear P-waves at all distances between $5^{\circ}$ and $15^{\circ}$, but between $15^{\circ}$ and $20^{\circ}$ the stations frequently report eP. A number of the times reported do not indicate any delay, although several stations report too late P. The twenty-degree discontinuity lies at $19^{\circ} - 21^{\circ}$ distance, depending on the location of the epicenters. For the earthquake of 1958, Jan 23, this distance was nearly $21^{\circ}$, which is abnormally great, depending on the unusually high velocity in the upper mantle.

A paper on this study has been presented as our Scientific Report No. 5 and has been submitted for publication in Annali di Geofisica.

## 3.4. Crustal Thickness in Fennoscandia from Phase Velocities of Rayleigh Waves by Eysteinn Tryggvason.

The phase velocities of Rayleigh waves traversing Fennoscandia, were determined by using all available records of short-period vertical seismometers in Denmark, Finland and Sweden. Two distant earthquakes were selected for the study, one with surface-wave propagation nearly perpendicular to the west coast of Norway, with epicenter in Mexico, July 28, 1957, another with wave propagation nearly parallel to the west coast o. Norway, with epicenter in the Kurile Islands, November 6, 1958.

The method demands, that each individual wave can be followed across the region studied. This involved some difficulties, because of large distances between the stations, and also because the waves were irregular and the wave fronts curved. Further difficulties were caused by the fact, that the direction of earth motion in Helsinki, Finland, was not known, and at Sodankylä, Finland, the direction of motion was opposite to that reported.

The irregularities in the form and direction of the wave fronts were chiefly caused by irregularities in the wave paths. As a consequence, each individual wave was recorded with different period at different stations. Furthermore, the curvature of the wave front was different for different waves in the same wave train.

All these facts make the result of this study less accurate, than those obtained by the use of more regular surface waves, as those originally used in similar studies (Press 1956). To get more reliable results, the curvature of the wave fronts was computed, and included in the phase velocity determinations. This procedure increased the accuracy of the result considerably.

On the other hand, no account was taken of the different phase shift at different stations, but its effect was studied, and found to account for less than 1 % of the apparent velocity in all cases, and was thus insignificant, compared with other errors.

The crustal thickness obtained was nearly constant over the area studied, about 35 km. There were however indications of slight increase in the thickness to the west and north, to 35-38 km as a mean for northern Sweden against 33-35 km in Finland and southern Sweden.

A paper on this research has been given as our Scientific Report No. 2 and will be published in Annali di Geofisica.

3.5. Deep Seismic Reflection Experiments at Kiruna by Markus Båth and Eysteinn Tryggvason.

In August, 1955, a series of specially arranged quarry blasts in the Kiruna iron ore mines was recorded with a refraction apparatus at approx. 10 km distance. The experiments are the first seismic investigations of the deeper crustal structure in Fennoscandia and were mainly undertaken in order to study near-vertical reflections from crustal discontinuities. The records show clear direct P waves with sharp onsets and a velocity of 5.65 ± 0.13 km/sec (in porphyry), furthermore S waves of less definite onsets and sound waves through the atmosphere. P waves reflected from crustal discontinuities are weak and of erratic occurrence, in agreement with theoretical expectation for near-vertical reflections. Approx. depths to the Conrad and the Mohorovičić discontinuities are calculated as 19 km and 33-34 km. There is general agreement with the explosion records written by the Grenet seismograph at Kiruna, and the differences which exist can be explained by the different frequency response of the two instruments.

A paper on this investigation has been sent as our Scientific Report No. 4 and will also be published in Geofisica pura e applicata.

3.6. Amplitudes of Explosion-Generated Seismic Waves by Markus Båth and Eysteinn Tryggvason.

Records of underwater explosions in Iceland in 1959 and 1960, mainly performed for crustal studies, have been investigated with regard to amplitudes. The amplitudes of P2 in the 1959 explosions and of the first arriving P waves in the 1960 explosions were found to be proportional to the first power of the charge weight, whereas amplitudes of P1 and S1 in the 1959 explosions were proportional to the 3/4-power of the charge up to 200 kg, but increased much slower for larger charges. The influence of water depth of shot point on the charge weights, required to obtain a certain amplitude, has been determined and it was found that the logarithm of the charge weight has a linear relation to water depth down to about 8 meters. The amplitudes of first arriving P waves decrease as the inverse 2.2-power of the distance up to about 30 km. For greater distances an exponential decrease of the form $(\text{const.}/\Delta)\, e^{-\beta\Delta}$ is valid. For P2 waves with a frequency of 10 cps we found $\beta = 0.027 \pm 0.003$ km$^{-1}$ for a profile across central Iceland and $\beta = 0.009 \pm 0.004$ km$^{-1}$ for a profile in the western part of Iceland.

The full paper constitutes our Scientific Report No. 6 and will be published in Geofisica pura e applicata.

## 4. Possible Continuation of the Present Work on Crustal and Upper Mantle Structure in the Scandinavia-Iceland Area

The border lines between the different structural regions, studied in this project, are not exactly defined. The crustal and upper mantle structure in Iceland and the Mid-Atlantic Ridge differs greatly from that found for Fennoscandia. We do not know, if these two structures are in contact along some border line, or if there exists a third type of crustal structure between the first two structures. This can be studied by using the refraction method at sea with artificial explosions in the Arctic-Atlantic Ocean off the Norwegian coast. Some information about this may possibly be obtained by using dispersion of surface waves of earthquakes in the Arctic-Atlantic Ocean, recorded at near stations, preferably at the Norwegian coast.

For more precise results, concerning the crustal structure in Fennoscandia, we need refraction measurements, using artificial explosions, along lines sufficiently long, to obtain the total thickness of the crust. Such measurements have been made in Sweden during the summer of 1960, but the result is not yet published. The refraction method using natural earthquakes may also give valuable information, and has been applied by Finnish seismologists for the northern part of Fennoscandia. This work should be continued and expanded, using the present net of seismograph stations with sensitive instruments.

The methods applying group velocities of surface waves are apparently of limited importance for studies of the crustal structure in Fennoscandia, because the local earthquakes are generally small and produce surface waves of very small amplitudes.

In Iceland the present investigation mainly concerns the thin lava layer, covering the whole island. The deeper crustal layers have been studied by using refracted waves of both natural earthquakes and artificial explosions (Tryggvason 1959, Båth 1960). Similar studies should be repeated in other locations to obtain a more complete picture of the earth's crust in Iceland. The extension and thickness of the "gabbro" layer below the ocean around Iceland can be studied by using the refraction method at sea, near the coasts of Iceland.

The thickness of the 7.4-layer, here identified as an upper mantle layer, can possibly be studied by large artificial explosions at sea, and with recording stations on land. Iceland is well located for such studies.

The phase velocity of surface waves in Iceland may possibly give valuable information about the deeper crustal structure of Iceland. However, this would require a more complete and homogeneous station net, than that existing at present. The great variation in crustal structure in the vicinity of Iceland may however make the application of this method difficult.

The phase velocity method can possibly be applied also on the whole Arctic-Atlantic Ocean north of Iceland and south of Spitsbergen, by using the existing net of stations in Greenland, Iceland, Norway and Sweden. It may however be necessary to add several stations in the region, preferably in Greenland, Bear Island and Jan Mayen Island, and standardize the instrumentation at the already existing stations, to make such a study possible. There are obviously great difficulties associated with such studies, due to large distances between the stations and great variations in crustal structure. Similar difficulties were partly eliminated in Fennoscandia, by including the curvature of the wave fronts into the velocity computation, but in case of the Arctic-Atlantic Ocean, these difficulties may be too grave for obtaining any reliable result.

## 5. Personnel

The following two seismologists have conducted the research described above, namely Dr. Markus Båth (Project Scientist), Seismological Institute, Uppsala University, Uppsala, Sweden, and Cand. real Eysteinn Tryggvason, on leave from Vedurstofan, Reykjavik, Iceland. Personnel at the Seismological Institute, Uppsala, has assisted in various parts, such as drawing of figures, typing of manuscripts, duplication etc.

Mr. Tryggvason worked full-time, i.e. at least 42 hours/week, on the research during the whole period. Dr. Markus Båth has worked part-time on the project, either as a co-author of papers together with Mr. Tryggvason, or by discussions and advice in the other papers.

The payment included in the contract has been used exclusively as salary to Mr. Tryggvason during his visit to the Seismological Institute, Uppsala.

References

Båth, M. 1958. Seismic Exploration of the Earth's Crust. Recent Developments. Geol. Fören. Stockholm. Förhandlingar 80:291-308.

Båth, M. 1960. Crustal Structure of Iceland, J. Geophys. Res. 65:1793-1807.

Båth, M. 1961 a. Die Conrad-Diskontinuität. Freiburger Forschungs-hefte C 101: 5-34.

Båth, M. 1961 b. Channel Waves in the Earth's Continental Crust. Scientia (in press).

Ewing, J. and M. Ewing, 1959. Seismic-refraction Measurements in the Atlantic Ocean Basins, in the Mediterranean Sea, on the Mid-Atlantic Ridge, and in the Norwegian Sea. Bull. Geol. Soc. Amer., 70:291-318.

Ewing, M. and F. Press 1956. Seismic Prospecting. Encyclopedia of Physics XLVII:153-168, Springer-Verlag, Berlin.

Gutenberg, B. 1944. Reflected and Minor Phases in Records of Near-by Earthquakes in Southern California. Bull. Seism. Soc. Amer., 34: 137-160.

Gutenberg, B. 1959. Physics of the Earth's Interior, Academic Press, New York.

Haskell, N.A. 1953. The Dispersion of Surface Waves in Multilayered Media, Bull. Seism. Soc. Amer., 43:17-34.

Kanai, K. 1951. On the Group Velocity of Dispersive Surface Waves, Bull. Earthquake Research Inst., 29:49-60.

Press, F. 1956. Determination of Crustal Structure from Phase Velocity of Rayleigh Waves, Part I: Southern California. Bull. Geol. Soc. Amer., 67:1647-1658.

Richter, C.F. 1958. Elementary Seismology, W.H. Freeman and Company, San Francisco.

Sawarenski, E.F. und D.P. Kirnos 1960. Elemente der Seismologie und Seismometrie, Akademie-Verlag, Berlin.

Steinhart, J.S. and R.P. Meyer 1961. Explosion Studies of Continental Structure, Carnegie Institution of Washington Publication 622.

Tryggvason, E. 1959. Longitudinal Wave Velocity in the Earth's Crust in Iceland. Natturufraedingurinn, 29:80-84.

UNCLASSIFIED

UNCLASSIFIED